Soil carbon storage responses to expanding pinyon–juniper populations in southern Utah

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Abstract. Over the past several decades, the expansion and thickening of woodlands in the western United States has caused a range of ecological changes. Woody expansion often leads to increases in soil organic matter (SOM) pools with implications for both biogeochemical cycling and ecological responses to management strategies aimed at restoration of rangeland ecosystems. Here we directly measure C and N stocks and use simple non-steady-state models to quantify the dynamics of soil C accumulation under and around trees of varied ages in southern Utah woodlands. In the two pinyon–juniper forests of Grand Staircase Escalante National Monument studied here, we found ~3 kg C/m^2 and ~0.12 kg N/m^2 larger C and N stocks in soils under pinyon canopies compared to interspace sites. These apparent increases in soil C and N stocks under woody plant species were dominated by elevated SOM in the surface 10 cm of soil, particularly within non-mineral-associated organic fractions. The most significant accumulation of C was in the >850 μm fraction, which had an estimated C residence time of <20 yr. Rates of carbon accumulation following pinyon–juniper expansion appear to be dominated by changes in this fast-cycling surface soil fraction. In contrast, we found that after separating >850 μm organic matter from the remaining light fraction (LF), C had residence times of ~400 yr and mineral-associated (MA) soil C had residence times of ~600 yr. As a result, we calculate that input rates to the LF and MA pools to be 10 ± 1 and 0.68 ± 0.15 g·m^{-2}·yr^{-1} (mean ± SE), respectively. These findings suggest that one consequence of management activities aimed at the reduction of pinyon–juniper biomass may be a relatively rapid loss of soil C and N pools associated with the >850 μm fraction. The temporal dynamics of the <850 μm pools suggest that carbon and nitrogen continue to accumulate in these fractions, albeit at very slow rates, and suggest that multidecadal storage of C following tree recruitment is limited to relatively small, subsurface fractions of the total soil C pool.

Key words: arid; carbon; Grand Staircase Escalante National Monument, Utah, USA; grazing; nitrogen; pinyon–juniper forest; soil; woody encroachment.

Introduction

The arid and semiarid regions of the intermountain western United States have experienced rapid ecological and human change over the past 100 years. These changes include a widespread expansion of woody plant species into adjacent plant communities (Archer et al. 1995, Van Auken 2000), changes in fire frequency and severity (Baker and Shinneman 2004), and widespread disturbance of soils through land use change (Neff et al. 2008). Arid and semiarid ecosystems are among the most disturbed places on earth (Asner and Martin 2004), and disturbance in these settings can lead to dramatic changes in ecological states (Westoby et al. 1989). In both the context of response to disturbance

and long-term sustainability of arid land ecosystems, soil organic matter (SOM) content plays a critical role. As the major repository for C, N, and available P, the organic pools are critical to the maintenance of nutrient cycles and aboveground productivity. In dryland ecosystems, these stocks of organic matter may be particularly important because these settings typically have relatively low concentrations of organic matter compared to more mesic environments (Batjes 1996). With small standing stocks of soil C and N, these ecosystems may be particularly vulnerable to disturbance (Fernandez et al. 2008), but conversely these soils may represent a potential target for activities designed to increase soil carbon storage (Lal 2004).

The response of arid ecosystems to the encroachment of woody species is highly variable, but typically involves a moderate to large increase in aboveground carbon stocks (Gill and Burke 1999, Asner et al. 2003, Hibbard et al. 2003, Hughes et al. 2006), although this is

not always the case (Schlesinger and Pilmanis 1998, Wilson and Thompson 2005). Changes in aboveground C inputs, whether positive or negative, also impact belowground C and N transformations (Gill and Burke 1999, Jackson et al. 2002, McCulley et al. 2004) and alter the spatial patterning of belowground resources (Schlesinger and Pilmanis 1998). Whereas the trend is for woody encroachment to increase aboveground carbon, the belowground response is substantially more variable. Soil organic matter stocks can increase, decrease, or stay the same after woody encroachment (Connin et al. 1997, Gill and Burke 1999, McCulley et al. 2004, Hughes et al. 2006). The implications of these changes are important because the broad expansion of woody biomass across the western United States and associated changes in carbon stocks may be large enough to impact the continental carbon balance (Birdsey et al. 2006). Estimates of the impact of western U.S. tree and shrub expansion on the U.S. carbon sink range from a negligible effect (Jackson et al. 2002) to a sink of 0.13 Pg C/yr (Pacala et al. 2001). Much of the uncertainty related to the size of this sink is due to the fate of belowground carbon pools following woody encroachment and there is an ongoing debate about the causes for this variation, with evidence that water availability may play an important role in the fate of belowground C following the expansion of woody biomass (Goodale and Davidson 2002, Guo and Gifford 2002, Jackson et al. 2002, Wheeler et al. 2007).

The majority of studies describing the effect of vegetation change on soil carbon stocks in drylands focus on changes in bulk soil carbon pools. Although important, the measurement of change in total soil carbon stocks provides only limited insight into the broader biogeochemical and ecological implications of changes in woody plant communities. Soil organic matter pools have a variety of turnover times, with a capacity for multidecadal carbon storage in only those fractions with slow turnover times. In many ecosystems, physically stabilized carbon pools will drive long-term carbon storage because mineral-stabilized forms of carbon turn over on decadal to century timescales (Trumbore 2000). As woody plant communities expand into grassland settings, the form and degree of stabilization of soil organic matter will play a major role in determining the long-term carbon and nutrient balance implications. If, for example, mineral-stabilized soil carbon stocks are responsible for the majority of soil organic matter changes associated with vegetation change, then the impact on carbon stocks will be long-lived (decades to centuries). If, on the other hand, soil carbon changes are dominantly in the non-mineral-stabilized organic fraction of soils, then these pools will likely be more ephemeral, prone to loss during disturbance, and more likely to respond to specific management activities. Additionally, the stabilization of N into these slower or faster turnover pools will have broad implications for the degree to which woody encroachment leads to increases in fertility under shrubs and trees (Schlesinger et al. 1996) and to the long-term implications of changes in soil nutrient content on restoration or management activities.

The role of physical stabilization in soil organic matter storage in dryland ecosystems is not particularly clear. In one study in a New Mexico desert, there was a relatively rapid incorporation of C from mesquite shrubs into the mineral-associated soil carbon pool (Connin et al. 1997). In other arid settings, however, the role of physically stabilized carbon is less apparent, with little evidence for correlations between soil silt and clay content and total soil C (Fernandez et al. 2008). The possibility of limited soil textural control on carbon suggests that mineralogic control of organic matter storage could be limited in some arid settings. These findings, in combination with evidence that soil CO_2 fluxes have nonlinear (e.g., threshold) responses to soil moisture and temperature (Fernandez et al. 2006), suggest that some arid ecosystems have relatively little stable soil organic matter and/or capacity for long-term carbon storage. Without additional evidence for the capacity of arid ecosystems to store carbon in long-term organic matter pools, it is difficult to evaluate the permanence of potential changes in soil organic matter associated with woody encroachment or to evaluate claims of a large carbon sequestration potential in dryland systems (Lal 2004).

The history of land use in the western United States also complicates interpretation of soil organic matter changes in response to tree and shrub expansion. Livestock grazing can have significant and long-lasting impacts on soil carbon stocks (Archer et al. 1995, Hiernaux et al. 1999, Neff et al. 2005) and may directly influence the expansion of woody species (Archer et al. 1995). Western rangelands have been extensively used for livestock grazing, and there is an exceptionally limited number of sites that have not experienced significant land use change during the 20th century (Van Pelt et al. 1992). In this study, we carried out an evaluation of carbon stocks under and between tree canopies in a site that has experienced very limited livestock grazing over the past century and a historically grazed site that has been grazed since the late 1800s in Grand Staircase Escalante National Monument, Utah, USA. Historical aerial photographs and stand age structure analysis from these sites suggest that there has been rapid recruitment in these populations over the last century (Harris et al. 2003; N. N. Barger, H. Adams, and C. Woodhouse, *unpublished manuscript*). We specifically focus on an evaluation of the rates of C and N accumulation under pinyon–juniper canopies to test the hypothesis that tree encroachment has led to increased soil carbon stocks and to evaluate the role of land use in mediating these responses. To further examine the nature of soil carbon, we also measured the distribution of organic matter through soil physical fractions to assess the degree of organic matter stabilization in these desert soils. These

measurements were combined with a non-steady-state modeling approach to evaluate the potential for these fractions to accumulate C and N in pinyon–juniper woodland soils following woody expansion.

METHODS

This study was conducted in Grand Staircase Escalante National Monument (GSENM) in southern Utah. This area of the Colorado Plateau contains a range of ecosystems including grasslands, shrublands, and pinyon–juniper woodlands (Stohlgren et al. 2005). Both sites receive ~360 mm of precipitation each year and are located at an elevation of 2000–2200 m. Mean daily minimum temperature at the site is −8.2°C and mean daily maximum temperature is 30.5°C. No Man's Mesa has not been grazed by cattle and was only grazed by goats for a brief period (two years) in the late 1920s. No Man's Mesa is geographically separated from surrounding areas by high cliffs and steep access trails. In contrast, Deer Spring Point has been extensively used for cattle grazing since the 1880s and is easily accessed from surrounding terrain. Soil for both sites is derived from eolian and alluvial Carmel-Page Formation sandstone. Soils are classified as a Pinepoint soil, which are well-drained sandy soils with sand content ranging from 85% to 90% across our sites. Vegetation structure at these sites can best be described as upland "wooded shrublands," sites where shrubs are dominant but support significant tree cover (see Romme et al. 2008; U.S. Department of Agriculture Natural Resources Conservation Service, *available online*).[6] Pinyon–juniper (P–J) wooded shrublands are intermediate to P–J savanna, which is characterized by a well-developed and often dense herbaceous understory, or old-growth P–J that maintains a dense P–J overstory with a depauperate herbaceous understory (Romme et al. 2008). The dominant shrub at this site was mountain big sagebrush (*Artemisia tridentata* subsp. vaseyana Nutt.(Rydb.) Beetle). Prior work at these two sites indicates that the sites have similar tree and shrub densities (Harris et al. 2003, Guenther et al. 2004), and related work indicates that tree ages are very similar between the two sites and that both sites underwent a large and synchronous expansion of pinyon cover in the time period since 1930 (N. N. Barger, H. Adams, and C. Woodhouse, *unpublished manuscript*). For this study, we controlled for slope angle by sampling on low-angle slopes (<15°). Field tests of soil CaCO$_3$ content was carried out with acid tests and observing soil effervescence. Sites were not located in areas where there was a strong indication of high CaCO$_3$ content.

Sampling techniques

We used bedrock geology and soil classification maps in a geographic information system (GIS) to establish

[6] ⟨http://www.ut.nrcs.usda.gov/technical/technology/range/mlra35.html⟩

six transects on similar soil and geologic settings on No Man's Mesa and six transects on Deer Spring Point. Transects were 300 m, with sampling sites placed every 25 m. At each transect stop, one tree core was extracted for dating tree ages. Each tree core was collected ~15 cm from the ground with an increment borer. We sampled soils at a midpoint between the tree trunk and the outer canopy edge and in an adjacent interspace without input of organic matter from trees. Two 10 cm deep soil cores were extracted at every point (one interspace and one under-canopy) with a volumetric soil corer (Soil Core Sampler Model 0200; Soil Moisture Equipment, Santa Barbara, California, USA). At every second stop, two additional cores, representing the 10–30 cm deep section of the profile, were extracted. At every third stop, we collected a 0–10 cm, >10–30 cm, and a >30–50 cm sample of the soil profile. With this sampling scheme, we collected 50 samples per transect: 26 with a depth of 0–10 cm, 16 with a depth of >10–30 cm, and 8 representing the >30–50 cm deep section of the soil profile.

The ecosystems in this area do not support a clearly separate litter layer above an O horizon. Rather, there is considerable intermixing of surface mineral soils and high organic content, particularly under tree canopies. In effect, the surface 10-cm samples include what would traditionally be the litter layer and the O horizon; in these ecosystems, the lack of soil layer differentiation makes it difficult to physically separate these layers, and so we used a variety of approaches to differentiate between soil C fractions.

Soil carbon analysis

All soil samples were weighed after oven drying at 60°C for 48 h. Particles larger than 2 mm were then removed by passing the sample through a 2-mm sieve that removed large-size plant materials. The >2-mm size fraction was, on average, ~2% of the mass of the sample and under 1% of total size carbon. Samples were then passed through an 850-μm sieve to obtain two size fractions; a <850-μm size fraction comprised of mineral and amorphous organic soil particles and a >850-μm size fraction comprised largely of recognizable organic particles. We chose the relatively large cutoff size because of the substantial contribution of relatively un-decomposed organic matter from under the tree canopies and because this large size cutoff allowed separation of a fraction that was entirely organic and clearly related to litter deposition. Each size fraction was weighed in order to back-calculate kilograms of C per square meter for each sample as a whole (<850-μm + >850-μm size fraction). To remove carbonates from the soil, each <850-μm size fraction was then split into a 2-g aliquot and acidified with 15% HCL until effervescence was no longer observed. After acidification, samples were re-dried and analyzed for C and N concentration with an EA 1110 CNS combustion analyzer (Thermo Electron, Waltham, Massachusetts, USA) at the Uni-

Table 1.   Estimates of organic and inorganic C stocks and organic N stocks for for pinyon–juniper ecosystems in the No-Man's Mesa and Deer Springs Point sites in Grand Staircase Escalante National Monument, Utah, USA.

C or N stock	Soil layer		
	0–10 cm	10–30 cm	30–50 cm
Soil organic C (kg/m^2)			
Canopy	5.32 (0.56)	1.59 (0.14)	1.25 (0.11)
Interspace	1.07 (0.11)	0.96 (0.09)	0.83 (0.07)
Soil organic N (kg/m^2)			
Canopy	0.23 (0.03)	0.12 (0.03)	0.08 (0.01)
Interspace	0.07 (0.01)	0.06 (0.005)	0.05 (0.004)
Soil CO$_3$-C (kg/m^2)			
Canopy	0.51 (0.25)	1.93 (1.38)	5.56 (3.66)
Interspace	1.13 (0.59)	4.96 (3.19)	5.02 (4.84)

Notes: We sampled soils at a midpoint between the tree trunk and the outer canopy edge and in an adjacent interspace without input of organic matter from trees. Values are means with SE in parentheses.

versity of Colorado, Boulder, Colorado, USA. All C and N concentrations are presented in this paper on a carbonate-free basis; however, estimates of C and N stocks include a correction of measured C and N concentrations to compensate for the removal of carbonates prior to analysis. In the majority of surface soils, this correction is very small (<1% of soil mass), but in some subsurface soils, there are somewhat higher concentrations of carbonates, and these values are presented in the results (Table 1). All surface soil samples (0–10 cm) were sampled volumetrically to allow calculation of bulk densities. Subsurface samples were not sampled volumetrically, and so for calculations of carbon stocks in subsurface horizons, we use an average of 1.6 g soil/cm^3 for mineral-dominated soils of low organic content in this area.

To further separate the type of organic matter in these soils, we carried out density separations on a selected subset of surface (0–10 cm), <850-μm samples from both the grazed and ungrazed sites for a total of 46 samples. Only sites located under tree canopies contained appreciable amounts of non-mineral organic matter, so for density separations we selected a range of samples from underneath trees with an age range of 50–400 years. The age range was used to examine the temporal dynamics of C accumulation under trees of varied age in these sites. For this analysis, pre- and post-density separation masses were recorded to determine the C and N fractional contribution from each density. The density separations were carried out in a sodium tungstate solution with a density of 1.68 to separate light (dominantly non-mineral) and heavy (dominantly mineral) soil fractions. These separations were repeated several times by centrifuging and decanting the suspended portion (light fraction) onto a Buchner funnel with a Whatman 50 filter. Each sample was centrifuged until all light-fraction material was removed and only heavy-fraction material remained at the bottom of the centrifuge tube. After centrifuging, each light and heavy

fraction was thoroughly rinsed with deionized water to remove residual sodium polytungstate, oven dried at 60°C, acidified to remove carbonate, and analyzed for C and N concentration. The >850-μm size fraction is referred to here as particulate organic matter (POM), whereas the <850-μm low-density fraction is called light-fraction (LF) organic matter, and the <850-μm high-density fraction is designated as mineral-associated (MA) organic matter.

Tree age dating

Tree cores were analyzed at the Institute for Arctic and Alpine Ecology, Dendrochronology Laboratory, University of Colorado, Boulder. Increment cores were mounted (Stokes and Smiley 1968) and progressively sanded with Federation of European Producers of Abrasives (FEPA) 120-, 220-, 320-, and 400-grit (100, 60, 30, and 20 μm, respectively) sandpaper using a Bosch belt sander (Robert Bosch, Farmington Hills, Michigan, USA). Cores were also hand-surfaced using 400- and 1200-grit sandpaper (20 and 15 μm, respectively). Cores were then visually cross-dated by using a previously developed pinyon pine (*Pinus edulis*) chronology for No Man's Mesa (C. Woodhouse, *unpublished data*). For samples difficult to visually cross-date, undated inner sections were run against a previously developed chronology using COFECHA (Holmes 1983, Grissino-Mayer 2001). All matching dates found using the COFECHA program were then visually verified. We recorded the first year of secondary growth to calculate tree age. Although we made three to five attempts in the field to obtain cores that included pith, we were not always successful. We estimated dates for cores without pith by overlaying sets of concentric circles on the inner rings of the core (Applequist 1958). To correct for the time to coring height, we collected pinyon seedlings from a range of sites and soil types across the region. Mean time to a coring height of 15 cm is ~11 yr. Once an inner ring date was obtained from the cores, 11 yr was

subtracted from this date to correct for the time to coring height.

Radiocarbon analysis

We measured the ^{14}C/^{12}C ratios in a representative set of soils from the LF and MA fractions of the <850-μm soil size class. These samples were separated and then freeze-dried on a Virtis BT6KEL-85 freeze dryer (SP Industries, Gardiner, New York, USA) fitted with an Edwards FL20K foreline trap (Edwards Limited, Crawley, West Sussex, UK) to prevent oil contamination of samples. The soils were analyzed for ^{14}C and ^{13}C at the Keck Accelerator Mass Spectroscopy facility at the University of California, Irvine, California, USA. All values are reported in Δ^{14}C and have reported uncertainties of <2‰.

Statistical analysis

It is not possible to replicate entire ungrazed mesa sites, so for our statistical comparisons, we used each 300-m transect as a replicate. We carried out site comparisons using multivariate analysis of variance (MANOVA) and regression analysis. For MANOVA tests, we used a mean value for the interspace or undercanopy and depth soil samples for an entire transect ($n \sim 13$ soil cores per transect) as a single measurement with a true replicate number equal to the number of transects on each mesa top (six on each). All soil samples were analyzed for C and N concentration, affording a full factorial MANOVA using C and N as response variables and site history (grazed vs. ungrazed), vegetation type (canopy vs. interspace), and soil depth as factors. For comparisons against tree age and other continuous variables, we use regression analysis. We also carried out an ANOVA of surface soil carbon and nitrogen stocks with site history, vegetation type, and depth as factors. All carbon turnover model fits were determined using iterative methods in JMP (SAS Institute, Cary, North Carolina, USA) and/or Microsoft Excel.

Quantifying soil organic C dynamics

The combination of C differences among interspace and canopy sites and variation in C stocks with tree age provide an opportunity to examine the non-steady-state dynamics of carbon accumulation in these ecosystems. The perturbation to soil carbon associated with tree recruitment at GSENM is still underway, and so the measurements made here (and in other similar settings) represent carbon stocks that are not in steady state. This aspect of soil carbon dynamics influences estimates of turnover time and accumulation rates, so here we present a non-steady-state approach to quantifying soil organic carbon turnover in the 0–10 cm soil fractions and outline a sequential procedure to estimate soil carbon changes through time. In all cases, we quantify changes in C stocks over "time," using canopy age (determined from tree ring analysis) as a surrogate for

the passage of hundreds of years. We first calculate the time dependence of aboveground C inputs to the soil by fitting a logistic model to the accumulation of the most ephemeral soil C pool (>850 μm). We then estimate turnover times by fitting the Δ^{14}C values and observed C accumulation in soil C fractions. Finally, we obtain an overall equation for the soil C fractions as a function of time by combining observed C accumulation data with the estimated turnover rate to fit the remaining unknown, the rate of C input to the fraction.

Our equations are based on a simple C balance for each fraction, i, of the form

$$\frac{dC_i}{dt} = I_i - k_i C_i \qquad (1)$$

where C_i is the C stock in the fraction (in grams per square meter), k_i is the turnover rate of the fraction, and I_i represents inputs to the pool. As written, k_i is consistent with all forms of C turnover, including loss as CO_2, dissolved organic carbon (DOC), or particulate transfer to subsoil horizons, and erosion. No attempt was made to model C stocks in the subsoil (>10 cm). Carbon inputs were assumed to follow a logistic curve, based on the observation of very low inputs until individual trees emerged from shrub canopies. This slow emergence is followed by a period of rapid canopy development and a subsequent phase in which inputs are assumed to remain at steady state. To obtain a quantitative estimate of these C inputs, we fitted the logistic equation, written as

$$\frac{dC_{\text{(POM)}}}{dt} = rC_{\text{(POM)}}\left(1 - \frac{C_{\text{(POM)}}}{C_{\text{S(POM)}}}\right) \qquad (2)$$

to the pattern of accumulation in the >850 μm C fraction ($C_{\text{(POM)}}$), yielding the following result:

$$C_{\text{(POM)}}(t) = \frac{C_{\text{S(POM)}}}{(1 - e^{-rt}) + \frac{1}{C_{0\text{(POM)}}}e^{-k_{\text{pom}}t}} \qquad (3)$$

where $C_{0\text{(POM)}}$ represents the initial >850-μm C stock (0.001 ± 0.0005 kg/m^2 [mean $\pm$ SE]), C_S is the steady-state >850-μm C stock (3.0 ± 1.2 kg/m^2), k_{pom} is a rate constant (0.13 ± 0.04 yr^{-1}), and t is the tree age as determined by dendrochronology (in years).

For the LF and MA, we estimated C dynamics using the following C accumulation equation, which is a solution to Eq 1:

$$C(\text{age}) = \frac{I}{k}\left(1 - e^{-k(\text{age})} + C_o\right) \qquad (4)$$

where I is carbon inputs (in grams of C per square meter per year) and age is tree age in years. In this equation I represents inputs entering the fraction and k is the turnover rate (per year). Turnover rates were estimated by fitting a Δ^{14}C model and are commonly expressed as turnover times ($1/k$). The Δ^{14}C model for estimating k and remaining parameters in Eq. 4 were fitted using an iterative
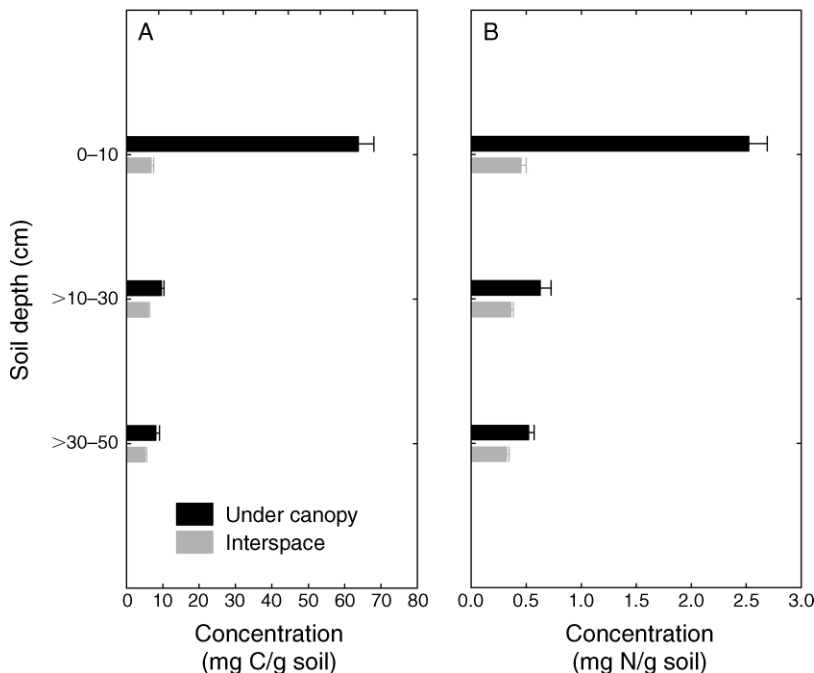
FIG. 1.   Distribution of carbon and nitrogen concentrations (mean + SE) with depth and across landscape (pinyon canopies vs. interspace sites) in two pinyon–juniper forests of Grand Staircase Escalante National Monument, Utah, USA. Concentrations are presented on a carbonate-free basis.

procedure, repeated several times until the adjustments were less than analytical uncertainty. The initial stocks C_{0LF} and C_{0MA} were estimated using the mean C stocks for those fractions in open (non-canopy) sites.

Changes in $\Delta^{14}C$ were modeled using a separate version of Eq. 1 established for ^{14}C, taking into account the variable $\Delta^{14}C$ of inputs driven by changes in atmospheric $\Delta^{14}CO_2$ as well as incorporating radioactive decay in losses. To implement this model of $\Delta^{14}C$ in LF and MA, we modified an existing annual-time-step model (Baisden et al. 2002) to account for non-steady-state conditions, specifically, the accumulation of C in the modeled pool following vegetation change. The atmospheric $\Delta^{14}C$ record was derived from Hua and Barbetti (2004) for 1955–1997, Baisden et al. (2002) prior to this period, and Levin and Kromer (2004) as well as Turnbull et al. (2007) for 1997–2005. All atmospheric $\Delta^{14}C$ data chosen reflect a spring–autumn growing season for mid-latitude regions of the Northern Hemisphere.

RESULTS

Organic matter concentrations and stocks

Soil carbon and nitrogen are generally low in the study sites in GSENM, with the exception of the surface soils under tree canopies. Grazing history had no impact on soil C or N stocks, but there were significant differences in C and N content of under-canopy and interspace sites ($F_{2,62} = 88.7$, $P < 0.001$), with surface under-canopy sites averaging 60 mg C/g soil and 2.5 mg N/g soil and

interspace sites ranging between 10 and 20 mg C/g soil and between 0.3 and 0.5 mg N/g soil on a carbonate-free mass basis (Fig. 1). Soil C and N concentrations also varied with soil depth ($F_{4,124} = 41.4$, $P < 0.001$), but with a more significant proportional reduction in concentrations under canopies than in the interspace (depth × canopy/interspace interaction term $F_{4,124} = 39.0$, $P < 0.001$; Fig. 1). Soil carbonate stocks are typically <10 mg C/g soil in surface soils and increase to an average of 30 mg C/g soil in subsurface soils.

The C and N concentration differences under tree canopies are driven by changes in the accumulation of organic matter under the trees. Overall, the accumulation of organic matter under trees is caused primarily by the accumulation of POM and LF organic matter (Fig. 2). The implication of this trend is evident in the significant negative exponential relationship between soil carbon content and soil bulk density ($P < 0.001$, $r = -0.8$, %C = $78.09 \times \exp(-3.05 \times$ bulk density)). High values of bulk density in the figure (>1.5 g/cm^3) are nearly exclusively mineral, whereas the values below 0.5 g/cm^3 (with organic carbon contents of ~10%) are in sites (nearly exclusively in under-canopy settings) where significant amounts of low-density organic material are intermixed with mineral soil.

Soil C and N stocks follow similar patterns to the organic matter concentration data above with higher under-canopy stocks of C and N compared to interspace sites ($F_{2,173} = 22.7$, $P < 0.001$), higher surface organic C and N stocks compared to subsurface soils ($F_{4,346} = 15.1$, $P < 0.001$) and no significant effect of site grazing
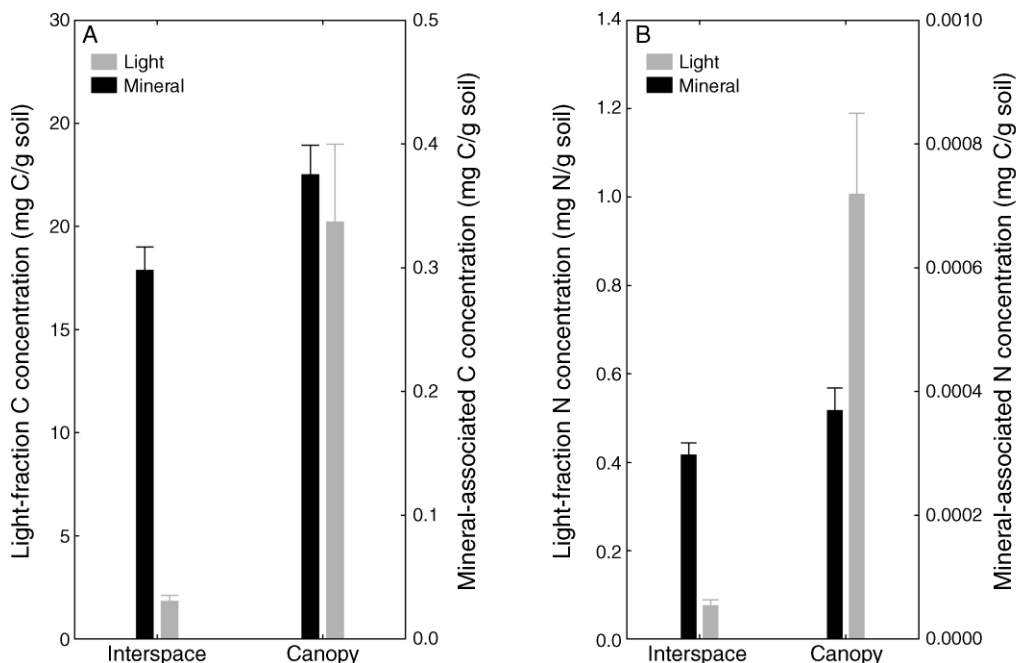
FIG. 2. Distributions of carbon and nitrogen in light (<1.7 g/cm^3) and mineral-associated (>1.7 g/cm^3) soil fractions. (A) Concentration of C in interspace and under-canopy soils in light and mineral-associated fractions. (B) Concentration of N for both density fractions in the interspace and under-canopy soils.

history (Table 1). Both the light- and heavy-fraction element pools reflect these patterns in both concentrations (Fig. 2) and in stocks.

Soil C dynamics

Parameters describing soil C dynamics, including turnover, were estimated by fitting the C stocks in each fraction vs. age to Eqs. 3 and 4 (Fig. 3). Fitting the POM fraction to Eq. 3 produces turnover (k_{POM}), $C_{0(POM)}$, and $C_{S(POM)}$ estimates (Table 2). The primary purpose of fitting Eq. 3 was to quantify changes in the rate of C inputs to the LF and MA pools as a function of tree age. As a result, the values for POM parameters reported in Table 2 should be viewed with some caution. In particular, the values reported for k_{POM} and $C_{S(POM)}$ imply a C input rate of 380 ± 190 g C·m^{-2}·yr^{-1}, which is considerably higher than a three-year estimate of litter-fall inputs of ~120 g C·m^{-2}·yr^{-1} from a pinyon–juniper ecosystem in Arizona (Grier et al. 1992). Factors that may explain this discrepancy include that the model implicitly includes belowground inputs (e.g., root litter), climate, and spatial variability, as well the fact that the use of Eq. 2 to estimate changes in input rates through time may oversimplify C dynamics. Additionally, these rates of input are for the area immediately under the tree canopy, and spatially integrated landscape-scale means would be considerably lower. Regardless, Fig. 3 shows that the $C_{0(POM)}$, and $C_{S(POM)}$, as well as our overall estimates of POM C stock vs. age, generate reasonable values given the large amount of variability in the data.

The soil POM accumulates rapidly in these sites and shows signs of reaching steady state (Fig. 3). In contrast, the dynamics of LF and MA accumulation appear much slower and benefit from the use of radiocarbon to constrain the apparent turnover rate. By modeling both total C and ^{14}C using Eq. 4 in an iterative procedure to constrain both inputs and turnover rate, we calculate the LF turnover rates shown in Table 3. In the six samples with measured Δ^{14}C, calculated residence times occupy a large range, spanning 237 years to 613 years, with an estimate of 433 ± 61 years. This residence time estimate is much longer than many previous estimates (Connin et al. 1997), reflecting the potential importance of correctly representing non-steady-state conditions in the model. The long residence times of LF reported here also demonstrate effectiveness of the 850-μm delineation used to separate POM prior to density separation and highlight the unusual dynamics of this fraction of soil carbon in dryland systems. The sample-to-sample variability in LF turnover estimates may reflect spatial variability, including variability in slow-turnover materials such as black carbon that may contribute to the fraction. The model fit suggests an input rate to the LF pool of 10 ± 1 g C·m^{-2}·yr^{-1}, indicating a relatively low C accumulation potential for this fraction when compared to site level net primary productivity (NPP).

Using the same procedure, we calculated MA soil C turnover time of 542 and 591 yr for two individual samples, yielding a mean estimate of 567 yr. The fit to Eq. 4 suggests a value for I of 0.68 ± 0.15 g C·m^{-2}·yr^{-1}, an order of magnitude lower than the estimated C input
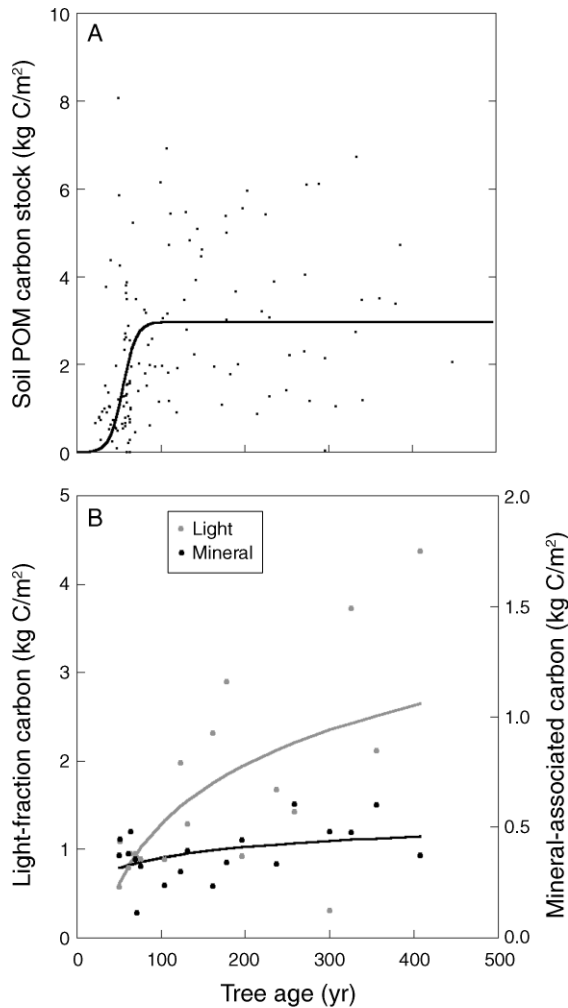
Fig. 3. Comparison of model fits and data for carbon in three soil fractions. (A) The modeled (solid line) and measured particulate organic matter (POM) fraction (>850-μm size class). (B) Fits to the light fraction and the mineral-associated fraction and data for these two fractions. Both graphs use tree age as a surrogate for time.

rate for LF. This I value implies that the accumulation rate of carbon into the physically protected MA forms in these ecosystems is <1% of NPP.

DISCUSSION

Pinyon pine recruitment in the 20th century in GSENM has resulted in significant increases in soil C and N storage. For carbon, soils under canopies have an average of ~3 kg of additional C than soils in interspace settings and for N the increase is ~0.1 kg. This accumulation happens largely over the course of the initial 100 years following tree establishment and translates into an overall soil accumulation rate of ~30 g $C \cdot m^{-2} \cdot yr^{-1}$ and 1 g $N \cdot m^{-2} \cdot yr^{-1}$, a number within the range of other studies of soil C accumulation following woody encroachment (Jackson et al. 2002). Despite this relatively large total rate of soil organic matter accumulation, the dynamics of the various organic matter pools are very different, with the highest rate of accumulation occurring in the fast-turnover POM pools, followed by far lower accumulation rates for the LF soil pools, and still slower accumulation into the physically protected MA pools.

Rates and forms of C accumulation

The majority of the carbon and nitrogen changes associated with pinyon recruitment into sagebrush communities occurred under tree canopies. This pattern of organic matter accumulation under trees relative to plant interspaces follows the "island of fertility" patterns observed in other desert systems and suggests a net gain of soil organic matter following tree recruitment in these ecosystems (Schlesinger et al. 1996), assuming that interspace soils are not becoming highly eroded, a pattern that was not observed at these sites. This pattern is in contrast to observations from other dryland ecosystems undergoing vegetation change that suggest a net depletion of soil organic matter following woody encroachment (Jackson et al. 2002). Taken in combination, these studies illustrate that there may be different trajectories for organic matter changes in dryland ecosystems undergoing changes in the woody component of these systems. Indeed, global syntheses of woody encroachment highlight the sensitivity and sign of soil C

TABLE 2. Parameter values describing dynamics of soil C fractions based on Eqs. 3 and 4.

Symbol	Description	Estimate	SE	Units
k_{POM}	POM turnover rate	0.13	0.04	yr^{-1}
$C_{0(POM)}$	initial >850 μm POM stock	0.001	0.0005	kg C/m^2
$C_{S(POM)}$	steady-state >850 μm POM stock	3.0	1.2	kg C/m^2
$1/k_{LF}$	LF residence time	433	61	yr
$C_{0(LF)}$	intial LF C stock	0.26	0.17	kg C/m^2
I_{LF}	input rate of C to LF	10	1	g C·m^{-2}·yr^{-1}
$1/k_{MA}$	MA residence time	567	NA	yr
$C_{0(MA)}$	initial MA C stock	0.29	NA	kg C/m^2
I_{MA}	input rate of C to MA	0.68	0.15	g C·m^{-2}·yr^{-1}

Note: Abbreviations are: POM, particulate organic matter; LF, light fraction; MA, mineral-associated fraction.

TABLE 3. Carbon isotope ratios, C stocks, and calculated mean residence times (MRT) from selected soil fraction samples.

Soil fraction sample, site, and age	δ^{13}C		Δ^{14}C		C stock at $t = 0$ (g C/m²)	Measured C stock (g C/m²)	MRT (yr)
	‰	SD	‰	SD			
Light fraction							
Deer Springs							
131 yr	−24.6	0.15	89.7	1.6	0.26	1.29	613
50 yr	−25.3	0.15	21.1	1.5	0.50	0.57	406
325 yr	−23.1	0.15	30.9	1.7	0.26	3.73	470
No Man's Mesa							
300 yr	−23.2	0.15	28.4	1.9	0.26	0.31	237
51 yr	−24.4	0.15	72.8	1.6	0.80	1.09	299
123 yr	−23.8	0.15	59.2	1.6	0.80	0.93	575
Mineral-associated fraction							
Deer Springs, 325 yr	−22.2	0.15	−10.3	1.5	0.29	0.48	542
No Man's Mesa, 300 yr	−22.4	0.15	−13.3	1.7	0.29	0.48	591

Notes: The samples span a range of tree ages. The soil fraction years refer to tree age where the sample was taken. We assumed an initial C stock and used measured carbon stocks and tree age as the input parameters for estimating mean residence time (MRT), as described with Eq. 4. The following values are point estimates for soil fraction turnover times under each tree where radiocarbon values were obtained: the residence time for the light fraction = 433 ± 149 yr (mean ± SD), and for the mineral-associated fraction = 566 (SD not determined; $n = 1$).

responses and the role that climate, soil properties, and land use history play in mediating these responses (Asner and Martin 2004, Asner and Archer 2008).

Despite net gains in soil organic matter beneath tree canopies, this study suggests that increases in organic matter occur largely in surface organic matter pools that are not stabilized physically. In this study, >80% of the carbon differences between canopy and interspace sites was due to the accumulation of surface non-mineral-stabilized C and N fractions. Although there is limited formation of a litter layer in these ecosystems, these large-size-fraction (POM) surface pools (generally holding more than half the soil carbon in these systems) clearly consist of recognizable plant detritus, with relatively little decomposed and humified SOM. Our estimates of turnover times in this surface organic matter pool yield mean residence times that are considerably less than 20 yr. This pattern of relatively rapid surface SOM turnover leads to a short period of C accumulation following tree recruitment with near steady-state conditions reached between 50 and 100 yr (Fig. 3A), a result consistent with rapidly increasing NPP paired with relatively rapid surface SOM turnover. Although the size of the surface C pool and its relation to tree age is admittedly quite variable, these data suggest that C is unlikely to be stabilized in this pool for more than two to three decades. The tree populations on GSENM have a median age of ~100 yr, and so these systems are likely near steady state with respect to surface POM soil carbon accumulation; accordingly, little additional stabilization into this pool would be expected in these mature pinyon–juniper forests.

The rapid turnover and low degree of stabilization of the large-size-fraction organic matter under tree canopies in these ecosystems may be a consequence of high surface decomposition rates in desert ecosystems. A study of decomposition dynamics near Canyonlands National Park found very rapid rates of soil respiration when environmental (temperature and moisture) conditions were adequate, conditions that occurred on average approximately one out of five days of the year (Fernandez et al. 2006). More generally, desert ecosystems are prone to rapid rates of decomposition and carbon release with pulsed resource availability (Huxman et al. 2004), and surface C is particularly vulnerable to loss due to a high degree of UV photo-oxidation of organic matter (Austin and Vivanco 2006). In the longer term, the relatively short residence time and lack of physical stabilization of the surface POM fraction makes the pool quite vulnerable to loss if either C inputs cease or environmental conditions change in a manner that favors increased decomposition.

The very rapid turnover of the surface POM fraction in this study stands in contrast to the comparative persistence of both LF and MA carbon in the soil subsurface. All carbon in the LF and MA fractions turns over very slowly, and even non-mineral-associated C appears to be relatively stable if it is intermixed with mineral soil. These fractions range from a small proportion of the total SOM in the first 100 years of tree growth to as much as half the soil carbon under very old trees. More-rapid turnover estimates have been presented for light-fraction SOM in the Jornada Experimental Range in New Mexico, USA (Connin et al. 1997), and for the Rio Grande Plains of Texas (Liao et al. 2006b), results that may be partially due to the transient nature of woody encroachment and the use of steady-state approaches to estimating turnover times in systems that are undergoing net C accumulation. As a point of comparison, a steady-state turnover time estimate based on stocks and input rates for these sites for the LF and MA fractions in this study would yield a turnover time of ~90 yr for the LF and ~600 yr for the MA fraction. This estimate is closer to the more-rapid

LF or POM turnover estimate from other studies. However, the evidence for continuing changes in the LF and MA small C fractions (<850 μm) of soils 300 yr after tree establishment suggest that steady-state assumptions for these fractions are not appropriate. Additionally, the separation of POM and LF in this study suggests that low-density organic matter fractions may exhibit a range of turnover times with some relatively recalcitrant components within the overall low-density fraction.

The potentially ephemeral nature of dryland soil C storage

The dominance of the light fraction, the non-mineral-stabilized SOM fraction, in carbon accumulation at GSENM has broad implications for carbon storage patterns in the western United States if the findings of this study are more broadly applicable. Work in Canyonlands National Park and in the Jornada desert (Connin et al. 1997, Fernandez et al. 2008) also suggests that most soil C is present in non-mineral or aggregate stabilized fractions, although studies in Texas show a greater proportion of stabilized woody biomass carbon in mineral fractions (Liao et al. 2006a). These variable patterns in the form of stabilized carbon are important to the long-term fate of soil fertility in these ecosystems and particularly to ecological response to management and disturbance. Environmental conditions or management activities such as thinning or rangeland improvement treatments that result in reduced NPP or the elimination of woody biomass could also trigger the loss of previously stabilized soil carbon and nitrogen if those fractions lack substantial physical protection. While soil carbon is often viewed as a potential multidecadal sink of carbon (Lal 2004), these results indicate that the majority of the SOM increase associated with 20th century tree recruitment in these sites is either in soil fractions with rapid turnover (e.g., the POM fraction) or in fractions that lack inherent physical protection mechanisms (e.g., the low-density fraction). Thus these pools could be prone to loss within two decades following disturbances that accelerate decomposition or oxidation. The extensive recent tree mortality in response to drought across the southwestern United States (Breshears et al. 2005) is one example of a broadly distributed change in vegetation dynamics that could precipitate large surface soil carbon losses in a relatively short time frame. Some pinyon–juniper woodlands have experienced up to 90% mortality in some stands. Such a dramatic alternation of forest structure over a few years may result in significant C release from these forests decades into the future. In addition to large-scale mortality, fires in pinyon–juniper-dominated ecosystems are another disturbance that may lead to relatively widespread loss of both aboveground and surface soil carbon. Soil carbon sequestration is a stated goal of the U.S. Healthy Forest Restoration Act of 2003, and so the ability to balance change in C stocks against other goals for ecosystem management will likely become an increasingly important issue for public land managers.

To provide long-term storage of soil carbon, non-mineral-associated carbon pools require some degree of physical or chemical protection. The accumulation of carbon in the LF and MA fractions is slow but illustrates some capacity of these soils to support additional carbon stabilization. The MA fraction, in particular, is a potential long-term storage pool, a point that contrasts with many agricultural ecosystems in which the mineral soil C fraction is relatively carbon saturated and unable to support additional C stabilization (Hassink 1997, Stewart et al. 2007). Once stabilized, the organic matter in the mineral fraction would be expected to have slow turnover times (Trumbore 2000), and this is the case for this study. Consequently, SOM age combined with the accumulation rates for both C and N shown in Fig. 3 suggest that this pool has the capacity to stabilize carbon on a multidecadal to century timescale but the accumulation rates ($\sim$1.4 g C·m^{-2}·yr^{-1} and 0.1 g N·m^{-2}·yr^{-1}) are low and represent a very small fraction ($\sim$5%) of the published estimates for total soil C accumulation in dryland systems (Jackson et al. 2002) following woody encroachment.

In addition to carbon stabilized in mineral pools, this study and others suggest that subsurface soil pools may have some additional capacity for multidecadal SOM stabilization. In Texas shrublands, there is some degree of aggregate protection of soil carbon pools (Liao et al. 2006a) that could confer longer term stability to SOM pools (at least in the absence of soil disturbance). Recent work in the Rio Grande Plains of Texas also suggests that woody and root-derived SOM may have increased chemical recalcitrance that may be responsible for some slowing of decomposition rates for this material (Filley et al. 2008). Structural (biochemical) inhibition of decomposition could be one potential reason for the long apparent MRT of the LF fraction in GSENM; however, further work is need to determine whether it is SOM chemistry or environmental conditions that limit the potential for decomposition in desert ecosystems. While these mineral and subsurface pools may provide a longer term sink for carbon, they also appear to play a fairly minor role in the soil C balance in GSENM ecosystems.

Grazing impacts on soil C storage

Grazing, like woody biomass changes, has a highly variable impact on soil carbon stocks. Our results indicate that grazing has had little to no influence on soil carbon storage in these sites, and this limited effect of grazing on soil carbon is in contrast to studies around Canyonlands National Park, where grazing causes large ($\sim$50%) decreases in soil carbon stocks (Neff et al. 2005, Fernandez et al. 2008). A number of the sites near Canyonlands and the GSENM sites examined here occur on Navajo Sandstones and have soils that are similar in texture, carbonate content, and soil nutrient

status. Most of the study areas in and near Canyonlands National Park, however, have been in grasslands with relatively little woody cover and an overall grass cover of >30% in contrast to the GSENM study area, where grass cover is ~5%. These two types of settings are typically represented as distinctly different ecological sites and are likely to have had different historical grazing intensity; these differences in historical grazing are one possible cause of variable responses. Sites within Grand Staircase are upland pinyon–juniper ecological sites, whereas sites within Canyonlands are generally Desert Sandy Loam (Indian Ricegrass) sites. Although historical grazing intensity is difficult to establish, these lowland grass-dominated sites likely experienced higher intensity livestock use because the vegetation is more suitable to cattle grazing. An alternative to the historical grazing hypothesis for variable SOM responses is the difference in environmental conditions at these sites. The Canyonlands area sites that are particularly prone to carbon and nitrogen loss are lower elevation sites that fall into the Semi-Desert category in the NRCS ecological site descriptions that generally experience lower annual precipitation (and higher evaporative potential) than the Upland sites. The role of historical grazing intensity vs. climate as a control on ecological responses to land use is an important question given concerns about the sustainability of dryland livestock operations (Brunson and Huntsinger 2008) and ongoing forest management/rangeland restoration activities across the Colorado Plateau.

Managing for ecosystem C in the future

The possibility of carbon storage in changing U.S. ecosystems has prompted increasing discussion of the possibilities of carbon management on public lands (Vine 2004, Meldahl and Kush 2006). Whether or not this is feasible depends to a large degree on the residence time and permanence of sequestered carbon pools. Additionally, any future management of ecosystems for carbon sequestration may come into conflict with existing federal lands policy that seeks to mitigate fire hazards (the healthy forests initiative) or, in the case of the western United States, policies that seek to reestablish rangelands following woody expansion. This study suggests that while there has likely been a significant multidecadal sequestration of C into soils following pinyon–juniper expansion in these ecosystems, the bulk of carbon stabilized in the process is present in short-residence-time pools. Thinning and overstory removal activities in pinyon–juniper ecosystems that seek to reduce hazardous fuel load should also be expected to cause relatively rapid declines in surface soil carbon and nitrogen storage. Similarly, the discussion about whether or not dryland ecosystems in the western United States are responsible for a substantial sink of carbon has centered largely on studies of the total soil carbon pool. Even if these ecosystems are storing substantial carbon in soils, the relatively large amount of carbon present in non-stabilized forms in soils at GSENM, Canyonlands, Rio Grande Plain, and the Jornada deserts (Connin et al. 1997, Neff et al. 2005, Fernandez et al. 2008, Filley et al. 2008) suggests that this carbon will be in forms not likely to be stabilized for more than a few decades.

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